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SYNOPTIC ANALYSIS AND FORECASTING OF SURFACE CURRENTS

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SYNOPTIC ANALYSIS AND FORECASTING OF SURFACE CURRENTS

Fleet Numerical Weather Facility
Monterey, California

Technical Note No. 9

June 1965

By

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and
T. Laevastu

FOREWORD

"Synoptic Analysis and Forecasting of Surface Currents" as reproduced herein is an edited version of the paper distributed previously in limited quantity as Fleet Numerical Weather Facility Technical Note No. 9.

This publication presents operational techniques for the analysis and forecasting of surface water currents from synoptic data. The difficult forecast problem has been reduced to manageable proportions by treating the total current as several distinct components at different time and space scales, and by describing the individual prediction relationships in terms of basic considerations and procedures.

Reviewed and approved on 22 June 1967.

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Captain, U. S. Navy
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TABLE OF CONTENTS

LIST OF FIGURES

INTRODUCTION

In navigation, fisheries and other sciences connected with the sea, it is often necessary to forecast or hindcast the direction and speed of surface currents at a specified time and place. Past attempts at current prediction have had only moderate success. It can be argued that progress on the quantitative understanding and prediction of currents has been hampered by a tendency to consider ocean currents to be analogous to wind currents in the atmosphere. Actually, it appears that ocean currents are affected by an even greater number of factors, and that different techniques may be required for successful prediction. This summary attempts to avoid past shortcomings by treating the total current in terms of a number of components. The existing quantitative knowledge of cause-effect relations of these components is first reviewed, and procedures are suggested for computation of currents in different time and space scales.

A summary of techniques for surface current prediction has been distilled from a voluminous literature search. Emphasis will be put on the basic considerations and procedures rather than attempting a mathematical reformulation of the problem. The theoretical treatises on large scale (oceanic to global) current depiction will not be discussed here; good references to these works can be found in textbooks and other summaries (e.g., Robinson [95]). Descriptions of some of the major ocean currents can be found in books by Stommel [109], and Reid [94].

This summary has been built upon an earlier draft by one of the authors; however, it has been extensively revised on the basis of recent experiences in analysis and forecasting of surface currents and their verification at the U. S. Fleet Numerical Weather Facility.

1. GENERAL NATURE AND VARIABILITY OF SURFACE CURRENTS; PRESENT STAGE OF KNOWLEDGE AND PREDICTABILITY

The literature on ocean currents contains a number of seemingly contradictory descriptions and theories. Many of the controversies arise from the differences which are observed in current behavior (both in space and time). These observed differences result in most cases from the various combinations of driving or influencing forces which prevail. For this reason, it is believed that ocean currents can best be understood and characterized by describing their space/time variability through study of their principal components.

Currents are typically unsteady in direction and speed (even in cases where the tidal components are subtracted and where the local winds vary little). This unsteadiness is well documented in the literature. Stommel [107], for example, investigated the surface currents off Bermuda, using drifting wireless-telemetering buoys. He found that there was considerable irregularity in current speed and direction, even during days when the wind was fairly steady. He calls this irregular motion, after Ekman, a kind of "macro-turbulence." Knauss [56] , using neutrally buoyant floats, also found "erratic" motion in deeper water below the thermocline (300 to 2140 m.). Reid [94] noted that short-period variations in the strength of the California Current have been observed of the same order as the strength of the flow (5-15 cm./sec.). The reasons for this variability have been sought in the condition that there are several other forces besides wind and tides which, independent of each other, affect the surface currents simultaneously. These are: (a) the "permanent flow" (either thermohaline or wind-driven large-scale circulation); (b) the changes of atmospheric pressure; (c) mass transport by waves; (d) internal waves; (e) the "modifying forces" (coriolis force, configuration of the bottom and coast, permanent and inertia currents); and (f) the periodic tidal currents. In addition, the conditions in neighboring areas (piling up, differences gradients in driving forces, etc.) affect surface currents in a given area and cause various eddies (both macro and meso scale) as well as convergences and divergences.

Various approaches have been used in the past for prognostication of surface currents, but actual synoptic analyses and forecasts have been very limited (with the exception of tidal currents).

In this manual an attempt is made to present those facts sifted from the literature and from personal observation which have application in analysis and forecasting of surface currents. The various components of these currents are evaluated separately. Suggestions are given for using these facts under field conditions, and results of large scale, synoptic current calculations by electronic computer are shown. In short, this manual does not pretend to be a complete summary on ocean currents as such but is primarily aimed at operational current analysis and prediction on a synoptic time scale.

Climatological surface current charts are widely used for the estimation of current drift. These charts, constructed from data on ships' drifts, are available for nearly all ocean areas. Their accuracy and usefulness have,

however, severe limitations. The data presented in current atlases are useful as a rough estimate of the actual current at a given time and place. The reports upon which these atlases are based were obtained from ships' logbooks and are nothing more than the difference between dead-reckoning fixes (from ship's speed and direction) and navigational fixes (astronomical, Loran, visual, etc.).

Hela [38] found that if the current displacement determined by astronomical fix is less than 2 nautical miles per 24 hours, the value in an atlas is entirely unreliable. In most cases, little weight can be given to displacements of 5 nautical miles or less per 24 hours. James [51] came to an analogous conclusion; namely, that the surface current charts can be considered reliable and accurate as mean values if there are at least 5 observations from 3,600 square nautical miles and the drift is more than 5 miles per 24 hours. There is obviously a real need for synoptic current charts, particularly in offshore areas (the data in some coastal waters are more reliable).

Numerous detailed current measurements taken with meters of various types are now available for limited areas. These have been useful in building up our knowledge and theories on surface currents. Most of these measurements have been made relatively near the coast, and coastal and tidal effects limit their use for drawing conclusions on offshore currents. Many of the current measurement series are also too short for proper evaluation. Measurements of actual surface currents offshore are extremely difficult to make. Detailed discussion on current measurement and various meters used can be found in excellent summaries by Wiegel and Johnson [125] and Paquette [88].

Several theoretical discussions of surface currents in the oceans are also available. Most of the theoretical models are concerned with large scale circulation and fall, therefore outside the scope of the present manual. The reader is referred to summaries on large scale circulation in the ocean by Stommel [108], Bowden [10] and Robinson [95]. Theoretical background and some descriptive material on currents can be found in textbooks such as Defant [20], Sverdrup, Johnson and Fleming [111], Proudman [91], Stommel [109] and the "Bibliography on Waves, Currents and Swell" by the American Meterological Society [1].

It has been customary to treat currents according to either Lagrange's approach (following the path of a given fluid particle) or Euler's approach (describing flow characteristics at a given point). Neither of these is followed consistently in this manual. In most cases, current is considered as the flow of water with a defined average speed and direction in a given time interval (synoptic period). However, a distinction is sometimes made between "mean current" -- which is considered as a current set with given average speed and direction in a time interval greater than the synoptic period -- and "actual current" -- which is considered the current set (with average direction and speed) in a given spot and time (time interval being 24 hours or shorter).

Although the present stage of knowledge on ocean currents is not perfect, the available material permits the initiation of synoptic analyses and forecasts. These analyses and forecasts are required as inputs in other synoptic oceanographic analyses and forecasts as well as for direct application *per se*.

Although the variability of currents is considerable, attempts should be made to evaluate quantitatively the causes of these variations. This can best be done through forecasting and daily verification. Forecasts have little value if they cannot be verified by actual synoptic analyses. Due to recent developments in other fields of oceanographic analysis and prediction, both direct and indirect verification are now possible (see chapter 5). Through these verifications additional knowledge is gained and existing ideas refined. It is to be expected that the approaches described in this manual will undergo considerable modification in the future.

2. EVALUATION OF THE COMPONENTS OF SURFACE CURRENTS

The speed and direction of the surface current at a given point, time and depth below the surface (e.g., at a depth of 3 m.) W_{ptz} can be represented as the resultant of the following components:

$$W_{ptz} = G_c + W_c + W_m + P_i + P_t + A_c + I \qquad (1)$$

where:

- G_c - permanent flow (gradient current and/or "characteristic current");
- W_c - wind current (lately often erroneously called the "Ekman current");
- W_m - mass transport velocity by waves;
- P_i - periodic portion of inertia current;
- P_t - periodic portion of tidal current;
- A_c - current caused by changes of atmospheric pressure and sea level;
- I - the velocity and directional component, caused by influencing factors, such as changing depth of water, coriolis force and coast and current boundaries.

The direction and speed are considered to be specified for all components. The present quantitative knowledge of these constituent components, applicable to current predictions, will be evaluated separately in the following paragraphs.

2.1 Permanent Flow

Permanent flow is directly related to pressure gradients in the sea which usually result from density differences of surface water columns between different locations ("thermohaline circulation"). The permanent flow is maintained by the great permanent and relatively constant wind systems ("wind-driven circulation") which cause piling up of water masses and changes in sea level, resulting in pressure gradients. It is believed by some investigators that the permanent flow is in most areas an "inertia function" due to prevailing winds and is therefore frequently referred to as the "characteristic current" (Palmen [87]).

For any synoptic analysis and forecast of actual surface currents it is necessary to separate the relatively persistent, slowly changing (Carruthers [13] and Carruthers, Lawford and Veley [14]) permanent flow from other components of surface currents which may vary hourly and daily and which should be predicted using the direct, synoptic influence of their driving forces.

In subjective forecasting (small-area), the permanent flow should usually be estimated using separation methods and then be added to other components computed and/or estimated by the synoptic consideration of their driving forces. In large scale numerical (objective) forecasts the permanent flow can be computed; however, "separated" permanent flow should also be used to verify these computations.

The basic method for separation of permanent flow has been developed and tested by Palmen [87] and Hela [37]. They took the currents associated with a certain wind velocity, grouped them according to the directions of the winds

and currents and computed the resultant currents for each direction (see formulas for wind current in section 2.2). The permanent flow was obtained by subtracting the computed current from the measured current. This method of separating out the permanent flow yields in most cases the so-called "characteristic current," thus indicating that the total surface currents are mainly wind-driven.

In areas where the data on average monthly surface currents are not available or are unreliable because of too few observations, it is necessary to approximate the permanent flow using other information such as sea level data, hydrographic (oceanographic) station data and observed winds.

Formulas have been derived for computation of currents from a pair of hydrographic stations (utilizing the difference in the dynamic heights at the stations) and for computation of gradient flow in a σ_t surface.

The formula of Sandstrom and Helland-Hansen for computation of a frictionless, stationary, relative current component cross the section between two stations is:

$$W_1 - W_2 = \frac{10}{2\,\ell\,\omega\,\sin\phi}\,(\Delta D_A - \Delta D_B) \qquad (2)$$

(symbols, see appendix). The computation of ΔD (the dynamic height anomaly) from hydrographic data is described in H. O. Pub. 607 (U.S. Navy, Hydrographic Office [117]). The relative current is positive (directed away from the reader) if station A lies to the right of B and if ΔD_A is greater than ΔD_B .

Equation (2) cannot be used in its basic form for computations of gradient currents in shallow water. However, Groen [32] has described an approximate method for estimating slopes and currents in shallow water. He replaced the solid earth by a fictitious section of water in which the density distribution was determined by putting the inclination of the α_a lines along every horizontal (or isobaric) line equal to the value projected horizontally from the bottom line. The lines were then extended from the actual water into the fictitious water by reference to these inclinations.

Although the original hydrodynamic approach, equation (2), was developed for use between two quasi-synoptic hydrographic stations, its use has been widened (but its limitations often neglected).

Mapping of dynamic height topography in relation to a reference level (e.g., 1000 m.) is assumed to give a general picture of circulation if enough quasi-synoptic hydrographic stations are available from the area (see fig. 2.1) Approximate current speeds are obtained by measuring the distances between contours.

Montgomery [78] concluded that a logical method of deducing oceanic flow patterns (permanent flow) from temperature, salinity and oxygen observations is to chart these properties for a surface of constant potential density. His expression for the gradient flow in a σ_t surface is:

$$W = \frac{37}{\sin\phi}\;\frac{\Delta Hg}{\ell} \qquad (3)$$

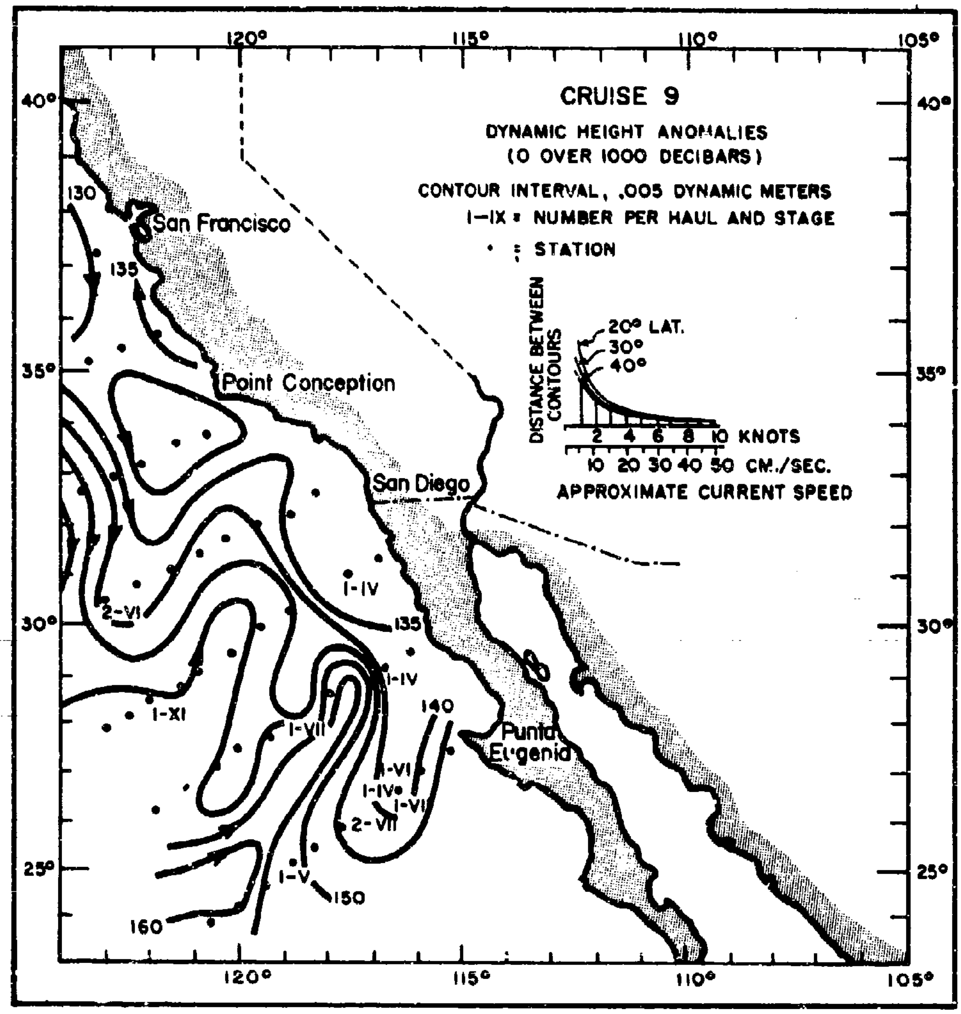

Figure 2.1 Example of a Chart of Dynamic Topography (after Johnson, 1960)

where Hg (the geostrophic potential) is:

$$\phi_a - \alpha_a p_n$$

The following just criticisms have been raised concerning the application of equations (2) and (3):

(a) The requirement that the stations in the section are quasi-synoptic is seldom fulfilled.
(b) The computation of ΔD requires an assumption of a "level or layer of no motion," which might exist but, at best, is difficult to assess correctly (several theoretical approaches for estimation of this layer exist, e.g., Defant [20]). This layer is generally assumed to be between 500 and 2000 meters. The results obtained from equation (2) also depend upon the spacing of stations.
(c) The equation cannot be used near the equator. (A review of the calculation of ocean currents at the equator has been made by Arthur [3]).

Some of the gradient currents, such as the equation counter-current in the central Pacific, are remarkably predictable in latitude and general physical characteristics (Austin, Stroup and Rinkel [4]). However, in most areas dynamical calculations with equations (2) give only a rough idea of the currents. Saelen [98] has made extensive comparison of geostrophic, computed and measured currents in the Norwegian Sea. Often he found only a qualitative agreement. Saelen (op. cit.) furthermore found that the conception of the circulation of surface and deep waters in the Norwegian Sea as a slow and regular drift does not hold, and currents at all levels vary rapidly over short periods of time.

As hydrographic stations are seldom synoptic, tidal fluctuations may lead to considerable errors in distribution of dynamic heights. Defant [19] has offered a method for elimination of "tidal" errors by preparation of dynamic charts for basins with simple tidal circulations, and Zaitsev [133] suggested a modification of this method for seas with complex tidal circulation, utilizing the co-tidal and co-range lines with coastal tidal data. For these corrections the following data are needed: (a) hydrographic station data; (b) co-tidal chart (usually theoretically computed M_2 component) of the area where co-range lines are indicated; (c) tidal curve for the period of the hydrograhic stations from one of the closest unsheltered representative coastal tide-gauge stations (reference station). The suggested correction procedure for an individual station is -- "The actual station time is noted on the tidal amplitude curve of the reference station. The co-tidal difference between the hydrographic station and the reference station is estimated from the co-tidal chart, and the time difference is transferred to the amplitude curve of the reference station (either to the right or left of the noted actual station time). From this point the tidal amplitude is read from the reference station curve. This amplitude is corrected with the ratio of amplitude differences between reference station and hydrographic station, and the resultant amplitude is used to correct the actual sampling depths --." This method does not account for internal tidal effects, which are thought to be considerable (Fjeldstad [26]) but have not yet been evaluated quantitatively.

Worthington [129] found that dynamic computations of the Gulf Stream, using closely spaced stations, indicate that the cross-current slope of the isobaric surfaces is uneven, resulting theoretically in three or four zones of high velocity separated by zones of relatively lower velocity. He suggested that the unevenness of the slopes of the isobaric surface is related to the surface temperature discontinuities observed in the Gulf Stream by von Arx and Richardson (1953), and that the stream is composed of overlapping, discontinuous currents at all levels. If this suggestion is correct, it would be virtually impossible to determine a "level of no motion" and apply hydrodynamical methods to estimation of currents in this region. (see also fig. 2.2).

In view of the limitations described so far in this chapter, it becomes obvious that the conventional hydrodynamical approach is unsuitable for numerical synoptic analyses and forecasting of the permanent flow. The main drawbacks are the following:

(a) The dynamic method gives only a qualitative picture of the currents, which can vary considerably over short periods.
(b) It is nearly impossible to obtain synoptic density distribution; the only good approximation must come from synoptic thermal structure analyses.
(c) The dynamic computations give the total current, but only the permanent flow component is desired, since other components seem to be more easily attacked by data on their driving forces.

It is therefore suggested that the best approach for numerical synoptic computation of permanent flow is to compute it from synoptic thermal and climatological salinity structure (tuned with appropriate constants) based on separation data at a number of localities. Such an approach is in use at the Fleet

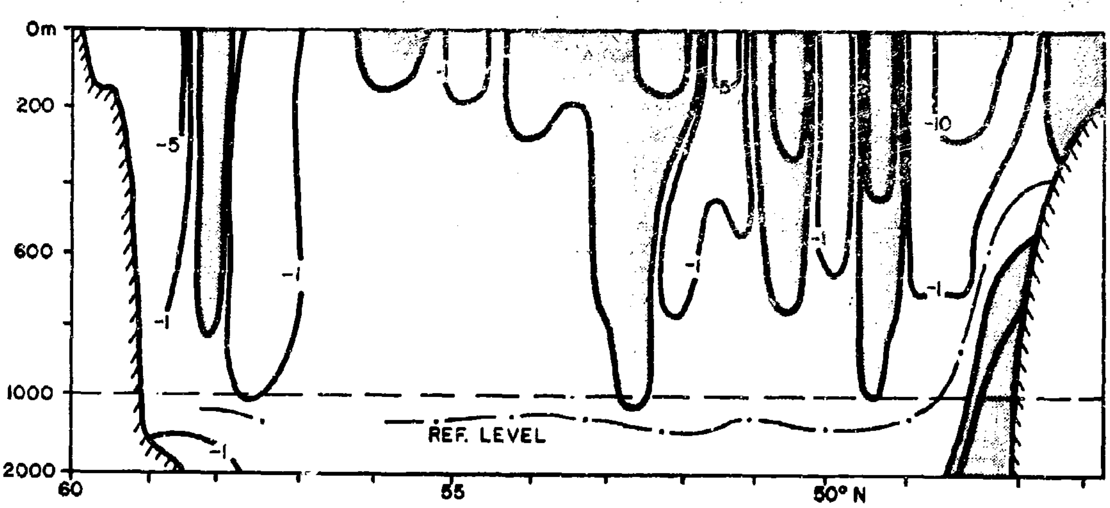

Figure 2.2 Distribution of Current Velocity (cm./sec.) Between Cape Farvel and Flemish Cape in Late Winter 1958. Positive Values Give Current to E or SE and Negative Values (gray areas) Give Current to W or NW. (Dynamical Computations; after Koslowski, [58] modified)

Numerical Weather Facility (Hubert [43]). Indirect support for this approach can be found in a few specific studies: Yasui [131] found that there is a close correlation between the distribution of temperature below the thermocline and dynamic depth anomalies. He constructed a simple empirical relationship by plotting temperatures versus computed dynamic depth anomalies, thereby shortening the dynamic computations. This relation, once established for a given location, has been assumed to be valid for a wider area.

Neglecting salinity, one can apply the well-known meteorological thermal wind relationship in the ocean if one knows the mean temperature of the layer between the surface and some deeper level of slow current velocity (selected as a level with negligible annual temperature change). The characteristic surface current is then given by:

$$W = \frac{g\Delta z}{f\bar{T}} \nabla \bar{T} \times |K \qquad (4)$$

where g is gravitational acceleration, f is the coriolis parameter, $\bar{T}$ is the mean temperature above the level of assumed zero current, Δz is the depth to zero current and $|K$ is the unit vertical vector.

The determination of a representative mean temperature ($\bar{T}$) is, of course, a critical factor in this problem. The temperature structure of the ocean is certainly not constant, particularly closer to the surface; therefore, synoptic temperature fields should be used if possible. The only place where sufficient data are available for reliable analysis on a daily basis is at the surface. (Fleet Numerical Weather Facility Sea-Surface Temperature [SST] analyses.) In order to include a part of the deeper temperature structure, the SST field is combined with a climatological field at 200 meters depth to obtain:

$$\bar{T} = K_1 T_{sfc} + K_2 T_{200} \qquad (5)$$

Finally, this field can be modified empirically in areas where salinity considerations are known to be important (Oyashio, Greenland, Labrador currents). This in effect corrects the ocean temperatures for salinity much as the meteorologist corrects atmospheric temperatures for moisture content when he uses the concept of "virtual temperature." The justification of the use of 200 m. temperatures and details of the numerical methods are given in chapter 4.

Some additional hydrodynamic approaches for estimation of currents from sea level and/or interface slope are given in sections 2.3 and 2.4. These approaches are of use where synoptic sea level records are available.

2.2 Wind Currents and Mass Transport by Waves

The wind is one of the main driving forces of surface currents, and many current systems are mostly wind-driven. Wind is directly or indirectly one of the main causes of temperature fluctuations in the sea and thus also affects the permanent flow.

In numerous cases a simple linear relation between the wind and current velocities has been found and/or used in the past. The current speed has been

found to vary from 1.2% to 4% of surface wind speed. There is evidence that the factor decreases with increasing wind speed, an observation which has caused some confusion and led to much discussion concerning a possible "critical" wind velocity. There is also some evidence that fetch length and wind duration have an effect on wind-currents; however, these factors have not been properly evaluated.

A simple relation can be used to approximate wind drift from climatological data (monthly resultant wind speeds):

$$W = K_1 V \tag{6}$$

The factor K_1 varies between 0.014 and 0.025. The value of 0.02 has been found most appropriate for a rough estimation of the sum of average wind and "wave" current in the surface layer (at about 3 m. depth). Although Thorade (from Defant [20]) assumed K_1 to be a function of latitude, there is no solid theoretical reason or empirical evidence for this.

The formula of Witting [127], based on more than 5000 observations, has been generally found to present best the relation between the actual wind and actual current velocities:

$$W = K_2 \sqrt{V} \tag{7}$$

where W is current velocity (cm./sec.) and V is wind velocity (m./sec.). The factor K_2 varies slightly with depth (see section 3.1). As an average, the value of 3.8 can be used if the mass transport by waves ("wave current") is estimated separately and an average current speed of the mixed layer is required. (If mass transport by waves is included in this formula, the factor is between 4 and 5).

According to Ekman's [25] theory, the direction of the wind current at the surface is 45° to the right in the northern hemisphere, and the current speed decreases with depth while the deflection becomes larger until, at a "depth of frictional influence," the direction of the current is opposite to that at the surface. Although Ekman's theory has been useful in many instances, it has not been verified in detail in the field. Ekman's theory cannot be applied to practical situations, because the conditions under which the problem was solved do not exist in nature (Belinsky and Glagoleva, [8]).

Recent investigations indicate that the wind current in offshore areas in middle latitudes deflects in general less than 20° to the right of the wind. (Bowden, [1953] - 18°; Gaul [30] - 15°; Lisitzin [68] - 12°; Stommel [107] -20°; see also Marmer [72]). Furthermore, it has been found that the current deflection depends also on the wind speed (Marmer [72]). Therefore the empirical formula of Hela [37], derived for conditions in the Baltic Sea, merits further testing in different latitudes and in the open ocean:

$$\beta = 34° - 7.5° \sqrt{V} \tag{8}$$

where β is the deflection in degrees to the right of the wind, and V is the climatological mean wind speed (m./sec.).

If the wind is weak, other current components can prevail, and the current movement is "erratic" in relation to the wind. At lower wind velocities the angle of deflection is somewhat larger, about 25° to 30° in higher latitudes, but converges rapidly between wind speeds of 4 to 6 knots to about 20° (Shulekin).

Tentatively, the following deflection angles are suggested for winds of 6 knots:

Latitude	Deflection (degrees)
0 – 10	0
10 – 30	5 – 8
30 – 50	10 – 18
50 – 70	20
70 – 90	25

With lower winds the deflection angle should be increased by 4° to 12°.

Some observations of the behavior of surface currents in oceanic areas indicate that the wind current is not always determined by the wind prevailing over a given location but rather by moving wind fields (and especially by stronger winds) over a larger area creating a relatively complex picture.

The relation between sea level changes (caused by wind set-up) and current has been rather extensively studied by Palmen and Lisitzin. Their results might be of use for estimation of currents. A formula, based on Torricelli's theorem and extensively tested by Lisitzin (e.g., [69]), is:

$$\Delta H_w = \frac{3.2\ V^2}{d} \tag{9}$$

The static effects of the atmospheric pressure differences must be eliminated beforehand; the slope of the surface (ΔH_w) is given in cm. per 100 km., wind (V) in m./sec. and depth of water (d) in m.

Mass transport by waves has been quantitatively accounted for in very few earlier surface current hindcasting or forecasting attempts but has been indirectly (and even unconsciously) included in empirical wind current formulas. This inclusion is advisable also in the future because existing uncertainties in quantitative evaluations do not justify a separate computation. In certain cases, especially in nearshore problems (e.g., longshore currents and currents in reef channels, etc.), it is necessary to account separately for mass transport by waves and for particle velocities in waves at different depths below the surface.

Some laboratory experiments on the mass transport by waves and their theoretical treatment are available in the literature (e.g., Masch [73]; Longuet-Higgins [70]). Unfortunately, laboratory experiments in wave tanks (for the investigation of mass transport) do not reproduce conditions found in nature. The transport is quite different in the beginning of the experiment when the wave generator is started (forward transport on the surface only). After a relatively short time, a circulation, determined by the tank boundaries, is established in the wave tank (backward transport on the surface

in the case of a mechanical wave generator, vice versa by "blower" generator, forward transport at some intermediate depth or on the bottom, depending on the relations between the wave length and the depth of the water).

According to theory, Gerstner waves are trochoidal and without mass transport velocity; Stokes waves, however, have a transport velocity (Lamb [63]; Pershin and Voznessensky [89]; Sverdrup and Munk [112]). A theoretical formula for mass transport velocity is usually given in the following form:

$$W_m = \frac{\pi H_{1/3}}{T} \; e^{\frac{-2\pi z}{L_{1/3}}} \tag{10}$$

This formula can be used with certain assumptions:

$$L_{1/3} \approx 50\sqrt{H_{1/3}}\,; \qquad \hat{C} \approx 1.25\, L_{1/3} \tag{11}$$

Masch [73] found that equation (10) gave values too small for mass transport as compared to the results of laboratory experiments. He concluded that the following equation (12) described the mass transport

$$W_m = 4\pi^2 \left(\frac{H_w}{L_{1/3}}\right)^2 C_e \frac{-4\pi z}{L_{1/3}} \tag{12}$$

The mass transport by waves decreases rapidly with depth, and the effect on the drift of a ship depends on the ship's draft. There is, unfortunately, no verification of the theoretical treatment and laboratory measurements through observations at sea. In wave tanks, equation (12) accounted for both wind current and mass transport by waves. Further tests of this equation in the sea are necessary (see fig. 2.3).

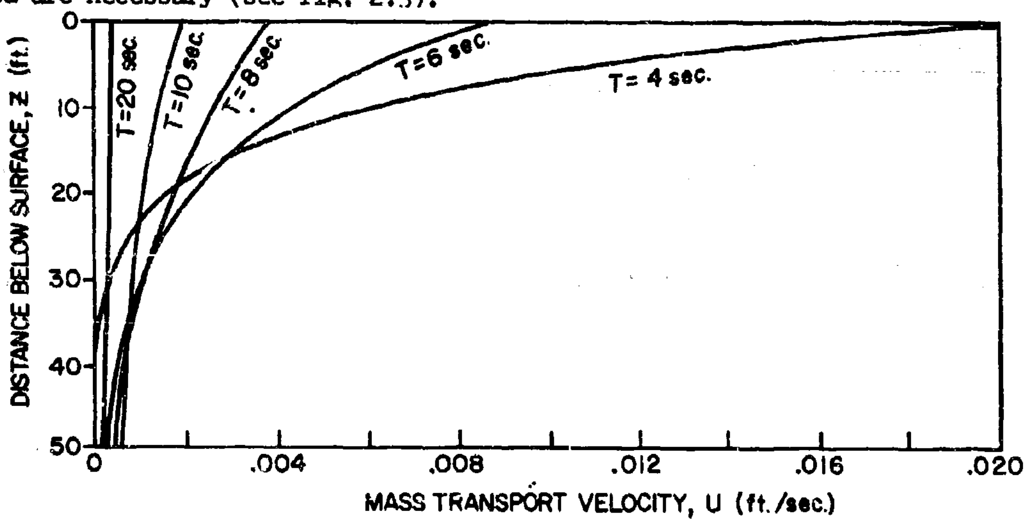

Figure 2.3 Velocity of Mass Transport at Various Depths as a Function of Wave Period (from Beach Erosion Board, 1942)

As the direction of mass transport is in most cases approximately the same as the wind current component, and as mass transport is of importance only close to the surface, practical surface current analyses and forecasts can ac-

count for this term by an empirical increase of the total effect.

2.3 Tidal and "Hydraulic" Currents

A knowledge of tidal currents is of great importance for navigation in coastal waters, especially in inlets. Therefore, predictions of tidal currents are made for a number of important locations. The interpolation between the prediction localities is, however, in many cases rather uncertain. Tidal current analysis and prediction for offshore waters have been greatly neglected, and it has been generally assumed that tidal effects are weak over deep water. Latest measurements, however, indicate that tidal currents can have considerable magnitude even in mid-ocean. For a 24-hour prediction the tidal currents might be neglected in most cases. However, in detailed interpolation of these forecasts, the tidal currents must be accounted for. Some specific aspects of tidal currents are pointed out below:

(a) The tidal currents are usually rotary offshore and reversing in coastal waters (see Hebard [36]). The tidal ellipses get narrower when one approaches the coast.

(b) In the northern hemisphere the sense of rotation is usually, but not always, dextral (clockwise).

(c) In general, the flooding current runs in the direction of the movement of the tidal wave (the direction can be ascertained from a co-tidal line chart).

(d) The rising tide is not everywhere synonymous with the flooding current, and falling tide does not necessarily mean ebbing current.

(e) High tide does not necessarily mean strong current, and a strong current may accompany a small tide. It is, therefore, always necessary to distinguish clearly between tide and tidal currents. The relations between the two are not simple and not the same in all locations.

(f) The behavior of tidal currents of otherwise similar tides is affected by the difference in location of amphidromic points of different constituents.

(g) Semidiurnal tide-producing forces cause proportional changes in tides and currents, but daily forces affect the current only half as much as the tides.

(h) The ebb usually lasts longer than the flood in coastal waters. How this can apply to rotary or nearly rotary currents offshore is not known.

(i) Some recent recordings of tidal currents in offshore waters indicate that the 14-day periodicity is a very prominent feature. (Mossby, personal communication.) This 14-day periodicity has been pointed out by Laevastu [60] in the fluctuations of thermocline depths as well as in fluctuations of sea level.

(j) The topography of the coastal area influences the tidal currents considerably (see, e.g., Neumann [83]) causing eddy formation behind headlands and in bays (Laevastu, Avery and Cox [61]).

(k) The tidal current ellipses are narrow and the currents strong in channels and on submarine slopes.

(l) The tidal current turns sooner in shallow water (close to the coast) than farther offshore.

An example of the behavior of tidal currents in relation to the tidal elevation curve in the Hawaiian Islands area is shown on figure 2.4.

In view of the description of tidal currents above, it can be concluded

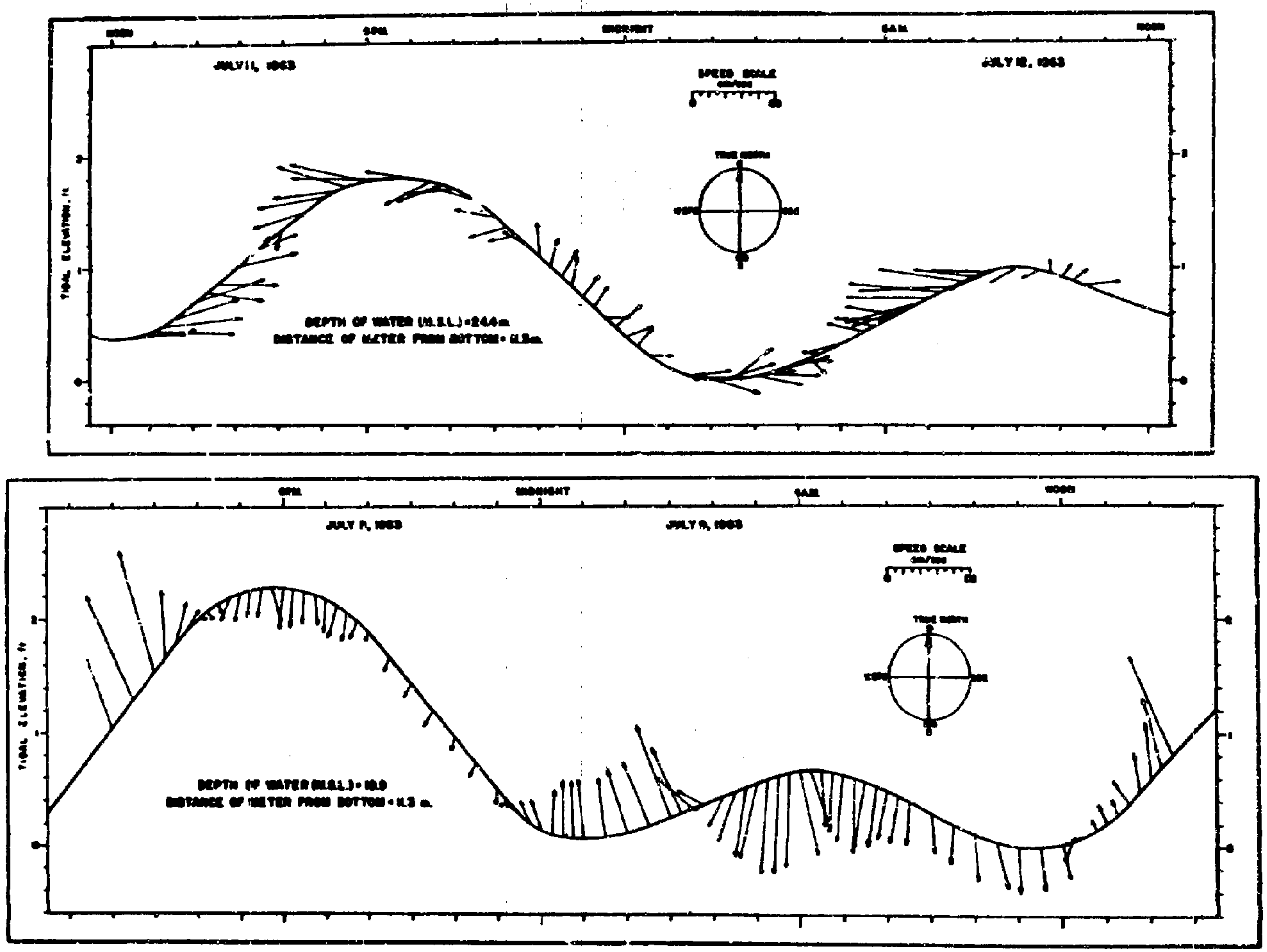

Figure 2.4 Examples of Recordings of Tidal Currents in Two Locations in Hawaiian Islands Area (from Laevastu, Avery and Cox [61])

that the only sufficiently accurate method for prediction of tidal currents in coastal waters is to record the tidal currents at a number of points along the coast for at least 25 hours (preferably 15 to 29 days), to analyze the changes of directions and speeds in relation to tidal elevation curves and then make predictions on the basis of these analyses. (See relation between tidal amplitude and currents, Marmer [72].) The currents between points of observations must be interpolated. The accuracy of this interpolation usually increases with decreasing distance between the points. If necessary and possible, other current components should be subtracted during the analyses and not included in the prediction of the tidal components themselves. Such predictions (based on empirical knowledge in a given location of the tidal currents in relation to tides) remain valid indefinitely.

At times harmonic analysis and prediction of currents can also be made. This is usually a standard procedure for the preparation of tidal current tables. The methods of harmonic analyses and prediction of tides and tidal currents are described in manuals such as U.S. Coast and Geodetic Survey (Schureman [101]), "Manual of Harmonic Analyses and Prediction of Tides" or in Admiralty of U.K. (Suthons [110]) "Admiralty Tidal Handbook."

In coastal waters with complex coastal topography and in narrows, many so-called "hydraulic effects" (differences in sea level) affect the tidal currents. The "hydraulic currents" in narrows are usually proportional to level differences and can be computed and/or predicted if the level differences at two locations and their changes with time are known. The "hydraulic currents" are known to exist in large areas in large archipelagos. If detailed current predictions in such areas are needed, these "hydraulic currents" can be accounted for with hydrodynamic computations, using sea level records. In field conditions and for occasional predictions, short-cut methods (utilizing sea level slope) have been used with success in the past.

If the atmospheric pressure were uniform and there were no "stationary" piling up of waters by wind action along the coast, the sea surface could be considered as an isobaric surface, and the gradient equation might be used to connect the inclination of the sea surface (tan i) to the surface current:

$$\tan i = \frac{2\,\omega \sin\phi}{g}\,\overline{W} \tag{13}$$

In the northern hemisphere, in the direction of the current, the isobaric surface would then slope upward to the right.

As shown by Sandstrom (1903), (from Hela [39]), there must exist, roughly, the following relationship between the transversal slope of the sea surface (i) and the mean current speed ($\overline{W}$) :

$$i = \frac{\Delta H_t}{B_c} = \frac{2\,\omega \sin\phi}{g}\,\overline{W} \tag{14}$$

In this formula, ΔH_t is the transversal height difference, and B_c is the breadth of the current. Hela [37] found the following empirical relation between the water height difference in cross currents and current speed in the

Strait of Florida:

$$\Delta H_t = 0.0152 W^2 \tag{15}$$

where ΔH_t is in cm. and W in nautical miles per day.

According to Torricelli's theorem the current speed can also be related to the longitudinal height difference (Hela [37]):

$$W_1 = \sqrt{2g \Delta H_\ell} \tag{16}$$

Margules' equation refers to the inclination of the interface (pycnocline) in relation to currents:

$$\tan \gamma = \frac{2 \omega \sin \phi (\rho_1 W_1 - \rho_2 W_2)}{g(\rho_2 - \rho_1)} \tag{17}$$

In a steady equilibrium state, the slope of the internal boundary surface in a two-layered sea will be greater than that of the physical sea level in the ratio:

$$\rho_1 : (\rho_2 - \rho_1) \tag{18}$$

The interface slope (i) should then be given by:

$$i = \frac{\tau}{g(\rho_2 - \rho_1) h_1} \tag{19}$$

Dietrich [22] has adopted the gradient current approach and the sloping interface relation for computation of currents in a lower layer along a sloping bottom with the following formula:

$$W = \sqrt{\frac{g h_2(\rho_2 - \rho_1)}{\chi \rho_2}} \quad \sin \alpha \tag{20}$$

The problems of prediction of tidal currents in offshore waters should be susceptible to numerical attack utilizing strictly hydrodynamic approaches. However, in the absence of such predictions, a subjective "field" procedure is suggested below (this approach requires verification).

If required, the tidal velocity component of a current in offshore waters could be approximately accounted for in short term forecasts (for example, for trawling purposes) using the tidal manuals for coastal areas and the co-tidal and co-range charts. The suggested simple, very tentative procedure for estimation of tidal currents in offshore areas is as follows:

(a) Assume that the tidal wave in offshore area is a progressive wave, which means that maximum current velocities occur at the crest and trough of the tidal amplitude curve for a given location, and about half of this velocity results between these points (i.e., a 1:2 tidal ellipse).

(b) Ascertain the stage of the tide and its amplitude in a given location and time from the co-tidal chart of M_2 tide and closest coastal reference sta-

tion along an open unprotected coast.

(c) Assume that the current is a function of depth

$$\frac{W_t}{\sqrt{h_1}}$$

where W_t is the tidal current at the reference station, and h_1 is the depth at any location. Multiply this result with the factor of

$$\frac{\sqrt{A_1}}{\sqrt{A_2}}$$

where A_1 and A_2 are the tidal ranges at the location in question and at the reference station respectively. (Assumption (c) is a very tentative one.)

(d) Assume that at the tidal wave crest the current runs in the direction of the movement (progress) of the tidal wave (direction from co-tidal chart), and at the trough it is opposite to the movement of the tidal wave (i.e., into the direction from which the wave came). Assume that the rotation of the current is clockwise.

It should be underlined that the above suggested procedure is largely hypothetical and not sufficiently tested.

Earlier attempts with different theoretical procedures indicate that the computed and observed velocities of tidal currents do not agree well in the open ocean. However, for a 24-hour prognostic current forecast for navigation, the tidal currents could be ignored because the resultant tidal transport for this period will be close to 0.

2.4 Inertia Currents and Other Current Components

When there is no apparent resultant force (except the deflecting force of the earth's rotation) acting upon a quantity of water in a given locality, the current is referred to as an inertia current. The primary force causing the inertia is usually the past prevailing and/or stronger winds in a given area, sometimes at some distance.

The periodic changes of this inertia current follow the theory (period and circle of inertia), although the forecasting of the phase at a given time and locality is nearly impossible at present.

The period of inertia is:

$$T_p = \frac{2\pi}{2\omega \sin\phi} \tag{21}$$

T_p is 12 hours at the poles, 24 hours at 30° latitude and infinity at the equator.

The radius of inertia is:

$$r = \frac{W}{2\omega \sin\phi} \tag{22}$$

Available investigations on inertia currents show that they affect the speed of the permanent flow at the given locality by periodic fluctuation. At present, no method exists for making a hindcast or forecast of these fluctuations; however, large and fast inertia eddies are often observed (e.g., Defant [17], Barkley, personal communication). It might be possible in the future to treat theoretically the inertia of a surface current system (created, for example, by a passing storm) and its response to new modifying forces (winds) by utilizing known relations of the moments of inertia and kinetic energy. It might also be possible to account for inertia currents (caused by strong wind fields) in continuous subjective analyses of large areas when more knowledge is accumulated from synoptic current analyses and forecasts and their verification.

Theoretically, it can be expected that in the open ocean an atmospheric pressure change will cause a wave in the pycnocline rather than a true surface current. However, there must be some current and mass transport connected with this wave. No direct empirical evidence is available on the current component caused by the change of atmospheric pressure. Recently, however, Donn and McGuiness [23] have shown that under certain conditions there can be a coupling of air pressure and ocean waves which can produce sea level oscillations 100 times greater than the barometric equilibrium wave. There is also the possibility that these "atmospheric pressure waves" are coupled with the inertia currents, especially when the inertia currents do not show dissipation with time (Defant [17]).

Internal waves, with tidal periods, may cause intermittent currents in the surface layers. The currents could be included in a prediction of tidal currents when accurate tidal current prediction techniques are available for offshore areas.

Some simple theoretical relations for the period and velocity of the propagation of internal waves are available for use in checking the importance of these waves in respect to currents:

$$t_i = \frac{2\ell}{\sqrt{\dfrac{g(\rho_1 - \rho_2)}{\dfrac{\rho_1}{h_1} + \dfrac{\rho_2}{h_2}}}} \tag{23}$$

$$W_i = \sqrt{g h_1 \frac{\Delta \rho}{\rho_1}} \tag{24}$$

At present, no tested method exists for predicting current component caused by the change of atmospheric pressure or by internal waves.

The current component caused by coriolis force does not appear in the formulas given earlier. This is, however, not strictly necessary, because the rational formulas are approximations and are usually based on experimental data which already indirectly include the coriolis component in the observations used. There is an exception in inertia currents where, theoretically, the

coriolis force is the only acting force. Coriolis force is otherwise a modifying factor and will be briefly considered in the next chapter.

3. EVALUATION OF MODIFYING FACTORS

The modifying factors of surface currents are here defined as the factors which do not directly cause a current but affect its speed and direction both in the horizontal and vertical. In addition, this chapter considers the change of current speed with depth in general and considers special relationships between surface and subsurface ("deep") currents.

3.1 Change of Current Velocity with Depth

Any current speed and direction change with depth may be related to internal friction, in connection with different currents in deeper water, bottom friction or viscosity. The change of current with depth can also be related to such features as density gradients with depth (stability), depth of the mixed layer and interactions between inertia and driving forces at the surface. No exact model of current speed change with depth exists.

Ekman [25] derived a theory for the change of current speed and direction with depth at the beginning of the century. This theory was later modified by Rossby and Montgomery, who related the deflection to latitude. Available measurements show that Ekman's theoretical consideration and the Rossby-Montgomery modification do not verify in the ocean, a fact recognized by Ekman himself. However, no available measurement series permits construction of a generally valid theoretical or empirical model of the current structure with depth, or a direct analysis of the factors involved, because of the shortness of the measurement series, lack of secondary data and great variability of existing data in space and time.

On the basis of theoretical considerations, the various components of surface currents differ considerably with depth. The pure tidal current should have uniform motion from the surface down (except very close to the bottom). The depth change of the wind current without friction should be a straight line with only a slight decrease of speed with depth. In an inclined wind drift with friction, the depth-speed graph is a parabola. The depth change of mass transport by waves is exponential. The permanent flow can be considered to have uniform speed with depth in the mixed layer but a relatively large speed gradient at the pycnocline.

Available current measurement series indicate that the current velocity change with depth in the surface mixed layer is variable in space and time. The greatest changes with depth occur near the surface, near the bottom and at the pycnocline. Some of these changes have been summarized by Stevenson [105]. His data have been utilized with the following assumptions and considerations in the construction of figure 3.1: (a) it is assumed that wind currents reach the depth of the thermocline; (b) the current speed falls to about 80% of its surface value at about 3 m. above the bottom or above a sharp pycnocline (average value of current at 3 meters from the surface being 100%). The upper curved part of the depth-speed graph (fig. 3.1) is caused by mass transport by waves.

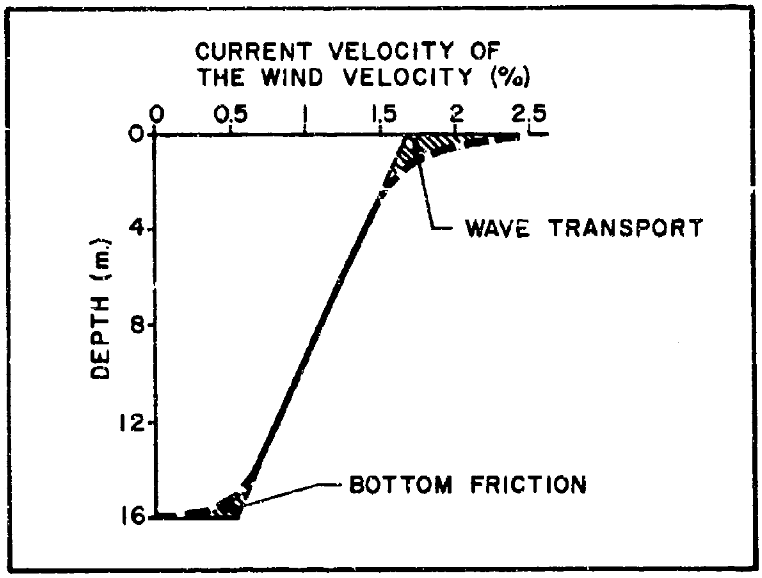

Figure 3.1 Approximate "Average" Change of Wind Current Speed with Depth in Shallow Water (after Stevenson [105], modified)

3.1 Coastal and Bathymetric Influences

The currents near the coast can be classified into two systems: (a) coastal current systems, consisting of relatively uniform drift, approximately parallel to the shore, composed of tidal, wind and gradient currents, and (b) nearshore current systems, consisting of coast-ward currents at the surface ("wave current"), a longshore current and seaward flow along the bottom and in the current rips.

The changing depth on the continental slope and shelf also influences current speed and direction. Fleming and Hagerty [28] found that, in general, the currents have a tendency to flow along depth contours where the depth changes more suddenly rather than following the configuration of the coastline. Usually the strong permanent currents flow along the continental slope.

Currents flowing along the coast with a slight "offshore angle" cause upwelling and intensive mixing off the slopes. Eddying and upwelling are also caused by steady currents along the coast if the direction of the coast changes suddenly (see fig. 3.2). A complicated configuration of the coast also complicates current patterns and requires detailed local evaluation. Islands in a steady current cause eddies on the leeward side and modify the current pattern at a considerable distance from the coast. In semiclosed bays, there is usually a "conformal current" along the coast, very similar to the currents observed in lakes. The tidal currents are usually the dominant currents on the continental slope. Upon these tidal currents there is a wind current component and near the estuaries a haline gradient component caused by runoff. True permanent flow can, in many cases, be neglected on the continental shelf.

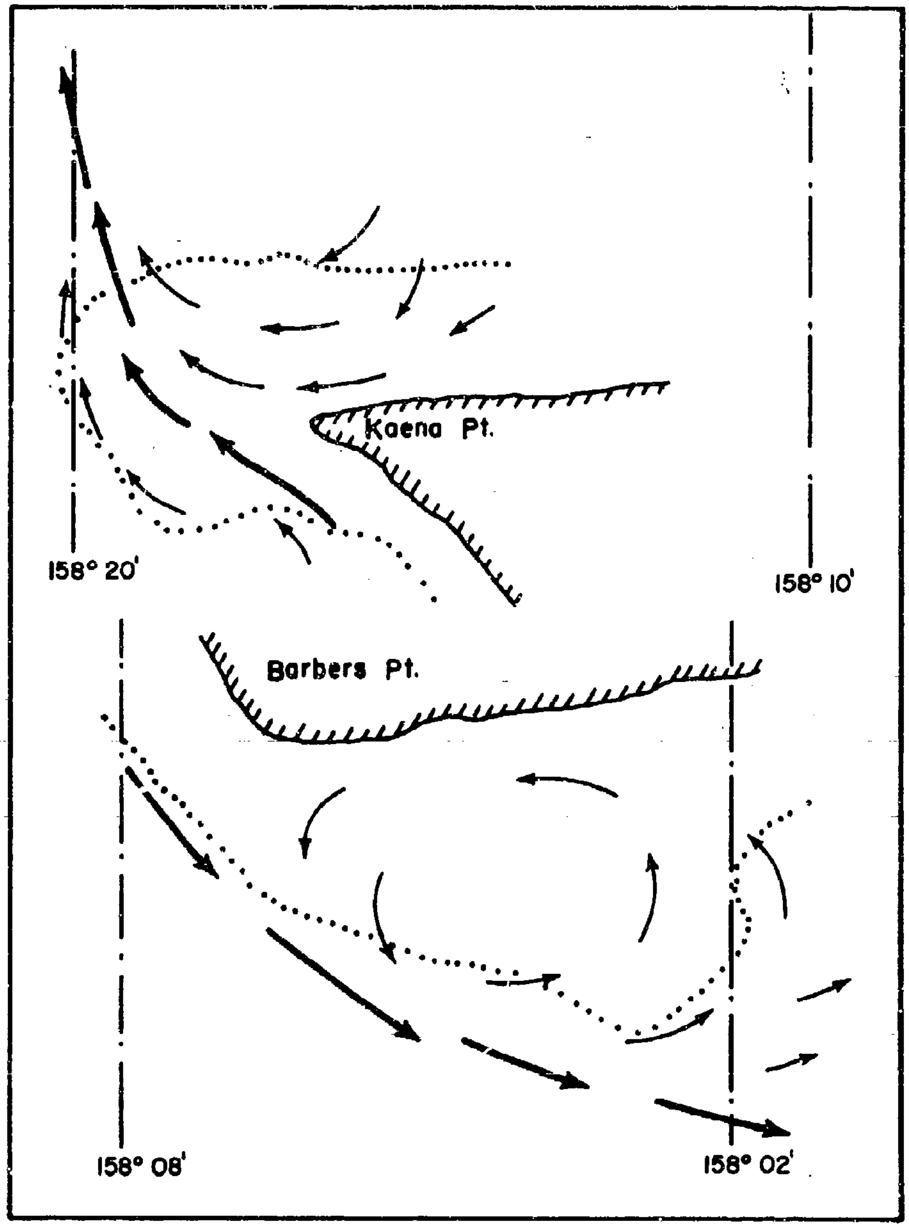

Figure 3.2 Examples of Eddy Formation Around Headlands (Generalized from Laevastu, Avery, and Cox [61])

The near-shore current system is illustrated in figure 3.3. Longshore currents can be approximately predicted along straight coasts (U.S. Navy Weather Research Facility [118]; Hunt [45]; Putnam, Munk and Traylor [92]. However, on irregular coasts, they depend on topography and are greatly influenced by sand bars, channels, etc. Figure 3.4 gives a nomograph for predicting longshore current. This nomograph is valid only for straight, uniformly sloping beaches with parallel contours.

Another formula has been proposed for prediction of longshore currents caused by mass transport by waves:

$$V = gmT \sin 2\,\theta_b \qquad (25)$$

where V is longshore current velocity, g, acceleration of gravity, m, beach slope, T, wave period and θ_b, breaker crest angle. No proper evaluation of this formula has been made.

No methods exist for predicting longshore currents on irregular and steep beaches. The local knowledge and experiences from similar locations and conditions forms the basis for any attempt to forecast near-shore currents (tidal component excluded).

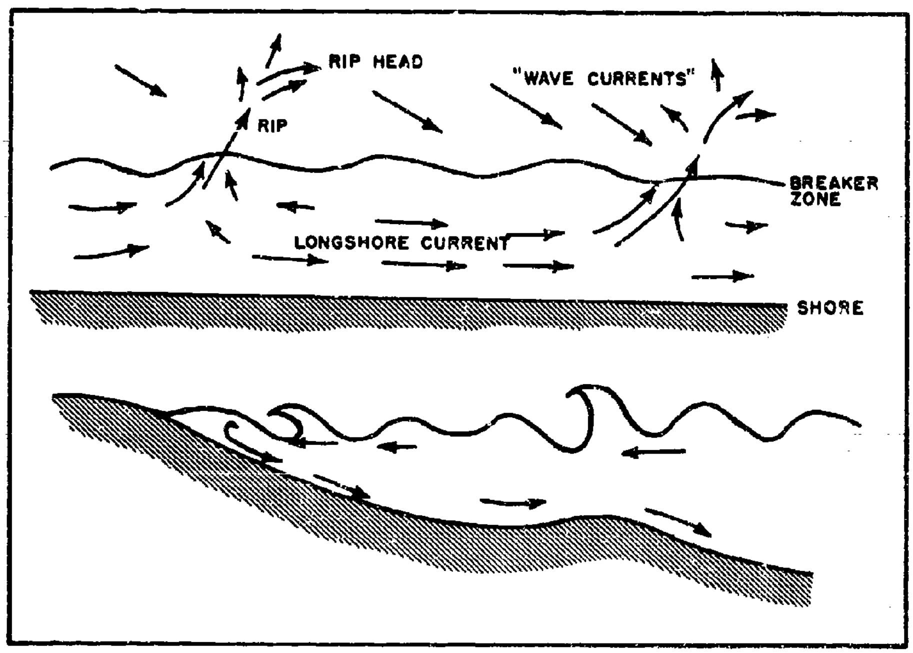

Figure 3.3 Scheme of Nearshore Current System

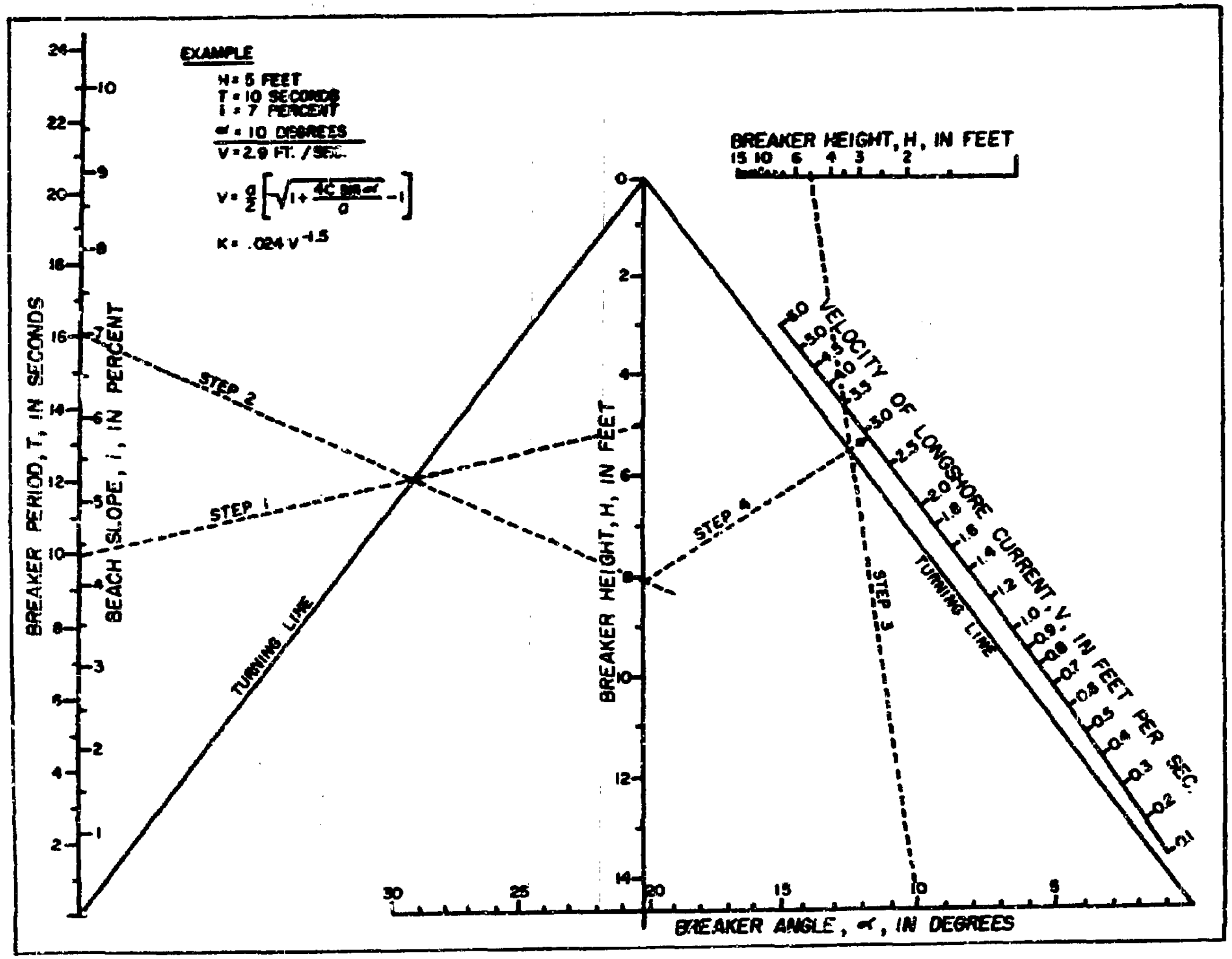

Figure 3.4 Nomogram for Determining the Velocity of the Longshore Current

3.3 Current Boundaries, Eddies and Deep Currents

The oceanic current boundaries and large-scale eddies associated with these boundaries have been described by several investigators - Uda [114], Fuglister and Worthington [29] and others. These current features must be taken into consideration in synoptic current analysis and prediction.

The types of current boundaries are: (a) dynamic (divergences or convergences of more or less permanent current systems); (b) topographic (caused by the topography of the bottom or coast); and (c) combined eddy systems. The most usual locations of the current boundaries are: (a) near meteorological fronts; (b) on continental slopes; and (c) around islands, capes and banks (local boundaries). Some minor surface convergences at continental slopes can also be caused by internal waves. The boundaries are frequently marked by: current rips, accumulation of flotsam (surface slicks), modified waves, roaring noises and fog. Usually there is also a change of water color at the boundaries. The surface temperature gradient on the boundaries can be 0.5° - 2° per 10 miles but may be even larger for major currents. McLellan [74] describes temperature changes off Nova Scotia of 3° C. per meter of depth and lateral changes of 1.5° C. per 100 meters.

Convergences of tidal current (tidal rips) can be rather extensive in some areas. Such tidal rips are often described by seamen; one such description taken from the Marine Observer (Vol. 27, No. 178, p. 203, 1957) is reproduced here:

"On the 7th of November 1956, between 0900-0930 S.M.T., the vessel sailed through a series of tide rips. These rips extended for about 10 miles on either side of the vessel in a 058° - 238° direction and gradually got progressively rougher, the final line having the appearance of a force 6 wind on the sea. Each line was about 100 yd. in width, with a 100 yd. stretch of smooth water in between lines. On passing through the final rip, the vessel swung 10° to starboard. The rips showed clearly on radar, and the P.P.I. screen had the appearance of a neatly ploughed field. There was no change in sea temperature. Air temperature 79° F., sea 80°. Light variable winds. Position of ship: 10 miles W. of Perlas Islands."

There exists a need for mapping the more permanent convergences and divergences of the oceans, based on ships' logs and on synoptic computation of convergences and divergences of the current fields.

The hydrographic and especially the dynamic conditions at current boundaries are usually rather complicated. Some graphical examples are given in figures 3.5 and 3.6.

The movement of current boundaries cannot be predicted subjectively at present with any great accuracy, and only certain rules of thumb can be established for estimation of their movements (Laevastu [60]). However, recent numerical synoptic anlayses provide several approaches which indicate their positions and gradients with considerable accuracy (Hubert, Slusser and Laevastu [44]).

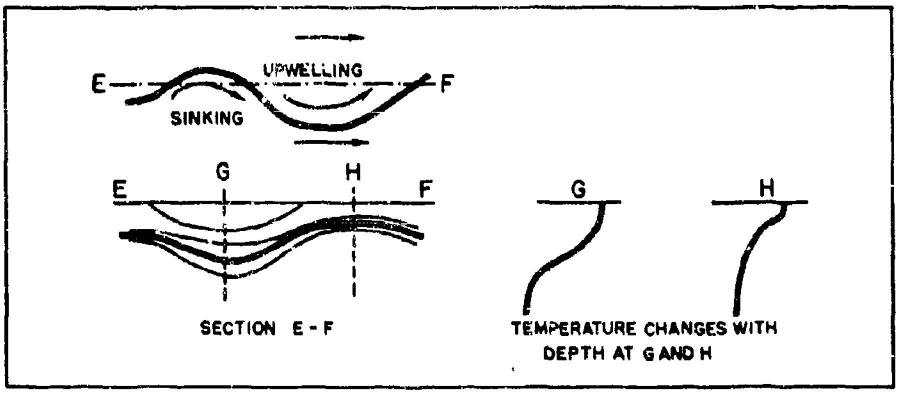

Figure 3.5 Scheme of Upwelling and Sinking and Resulting Subsurface Thermal Structure at a Meandering Boundary

The current boundaries, especially convergences, are associated with meanderings and eddies which cause sinking and/or upwelling of deeper water (figs. 3.5 and 3.6). The greater the speed difference on both sides of the boundary (and also the more irregular the bathymetry of the bottom) the more extensive is the meandering. Many of the large-scale meanderers are stationary, due to the topography of the continental slope. However their intensity seems to fluctuate, depending on changes in driving forces.

There are also great eddy systems which are caused by winds (cyclones and anti-cyclones). In general, the centers of cyclonic eddies (anti-clockwise) are cold, and they are associated with upwelling in the center. The centers of anti-cyclonic eddies are warm, and sinking occurs very slowly in the center. If a cyclonic eddy is cut off from a cold water mass into a warmer one, slow sinking takes places also in this eddy (fig. 3.7).

Several meteorologists and oceanographers, notably Palmen and Rossby, have described the similarities between atmospheric and oceanic structural-dynamic features. Newton [85] also found that oceanic and atmospheric eddies are comparable in the sense that in each case the maximum subsidence of the cold mass took place at about half the rate of horizontal flow of mass into the cold-water or cold-air tongue.

The currents in deeper layers of the sea are not subjects of this summary. However, some of their characteristics must be borne in mind when analyzing and forecasting surface currents inasmuch as they influence each other. The deep currents balance the "water budget" and bring to equilibrium the disturbances of the surface layers. Therefore, the deep currents may often be opposite in direction to those at the surface. The structure of deep currents with depth can be complicated (as shown by recent measurements with neutrally buoyant floats). Furthermore, these measurements indicate that the deep currents are relatively unsteady in direction and speed, contrary to earlier beliefs.

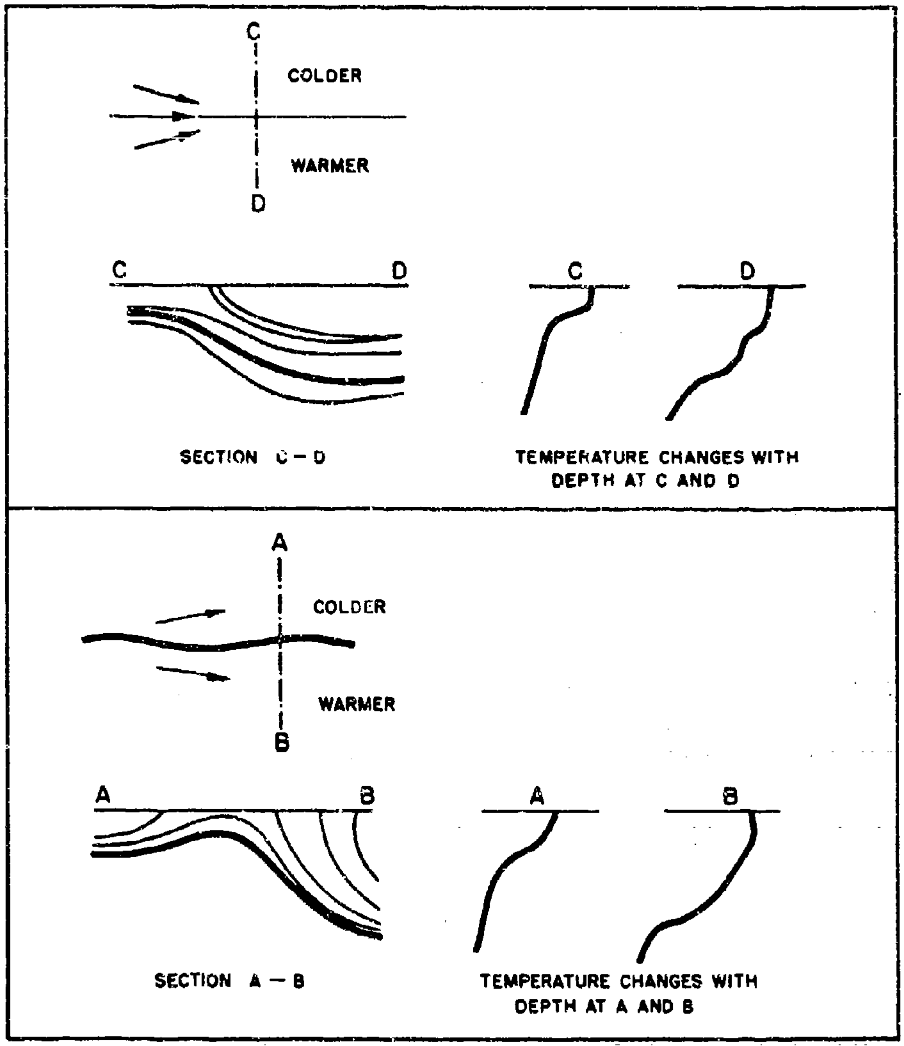

Figure 3.6 Structural Models of Convergences and Divergences

The well-known coriolis force as a modifying factor has not been mentioned earlier; it is usually indirectly included in the semi-empirical approaches of current analyses and forecasting.

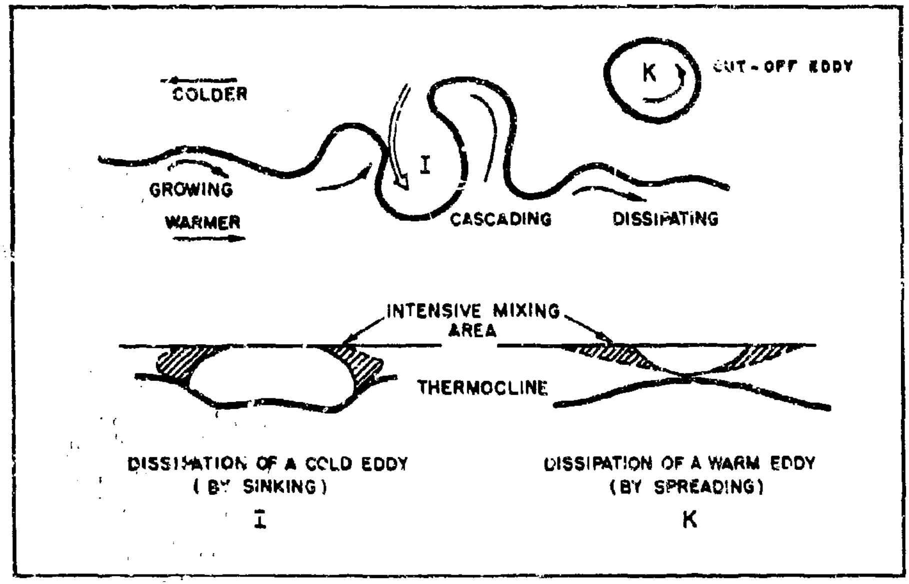

Figure 3.7 Mechanism of Cutting Off Eddies at Current Boundary and Their Dissipation

4. OBJECTIVE COMPUTER METHOD FOR ANALYSIS AND FORECASTING OF SURFACE CURRENTS

The detail and accuracy of analyses and forecasts depend greatly on the area and time scales used. This is true both for objective and subjective approaches. Small-scale analyses/predictions require one to program local knowledge of influencing factors as well as more detail of the driving forces. In this chapter, we describe briefly the principles of hemispheric (or oceanwide) analysis and prediction of surface currents on a 12-hour synoptic schedule.

The computation of permanent current is made with equations(4)and(5) (section 2.1). In recent programs the average temperature between 0 and 200 meters is found by integration, using FNWF standard levels of subsurface temperature analyses. As the analysis refers to surface currents, and as countercurrents are known to exist below the permanent pycnocline, there is no need to integrate the temperature field below 200 m. level. Salinity corrections are added on a seasonal basis.

The wind current component is computed with equation(7)and added to the permanent flow. Integrated true surface winds are used.

According to Ekman [25] , the direction of the wind current at the surface is 45° to the right of the wind in the northern hemisphere, and this angle increases with depth. Recent investigations reveal that the deflection is more nearly 12-20 degrees, being larger and more irregular at lower wind speeds (possibly because of the increased importance of other components) and smaller at higher wind speeds. As the surface wind is about the same angle to the left of the geostrophic wind, it can be assumed that the direction of the wind current is the direction of the geostrophic wind.

It is assumed that the current is relatively uniform and unidirectional in the turbulent mixed layer down to the thermocline (or about 200 meters). The mass transport due to waves, however, modifies this picture as this effect decreases exponentially with depth (Masch [73]). Therefore, if, in equation(7), V is in m./sec. and W in cm./sec., K_2 is taken to be 4.8 for surface currents (ship routing and drift computations) and 3.2 for the average current down to the thermocline (for convergence/divergence computations). Obviously, there is a time lag between the change of the wind and response of the sea. This lag seems to be shorter than previously believed and is partially minimized by the 24-hour averaging of wind speed. (see further, chapter 6).

Since all computations are carried out in the standard Fleet Numerical Weather Facility grid system, u and v current components are determined at approximately 200 nautical mile intervals for all northern hemisphere ocean areas. From these components, direction and total transport (nm./day) fields are determined and stored on magnetic tape for later output in chart form or as special messages giving the currents at specified latitude/longitude intersections. Some current transport charts are shown in figures 4.1 and 4.2.

The contours represent total current transport in nautical miles per day. One can clearly distinguish such well-known features as the Gulf Stream,

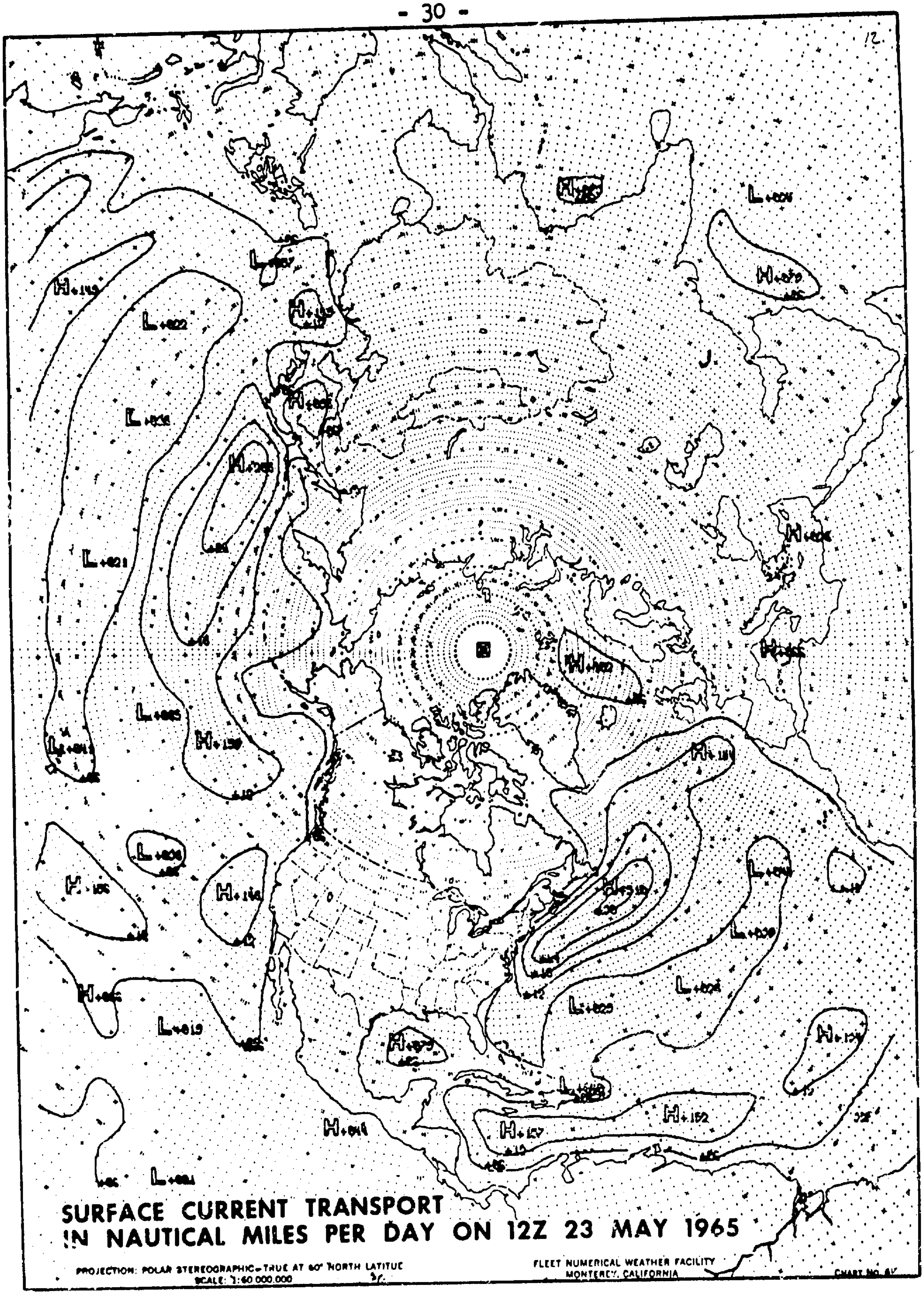

Figure 4.1 Surface Current Transport in Nautical Miles Per Day on 12Z 23 May 1965

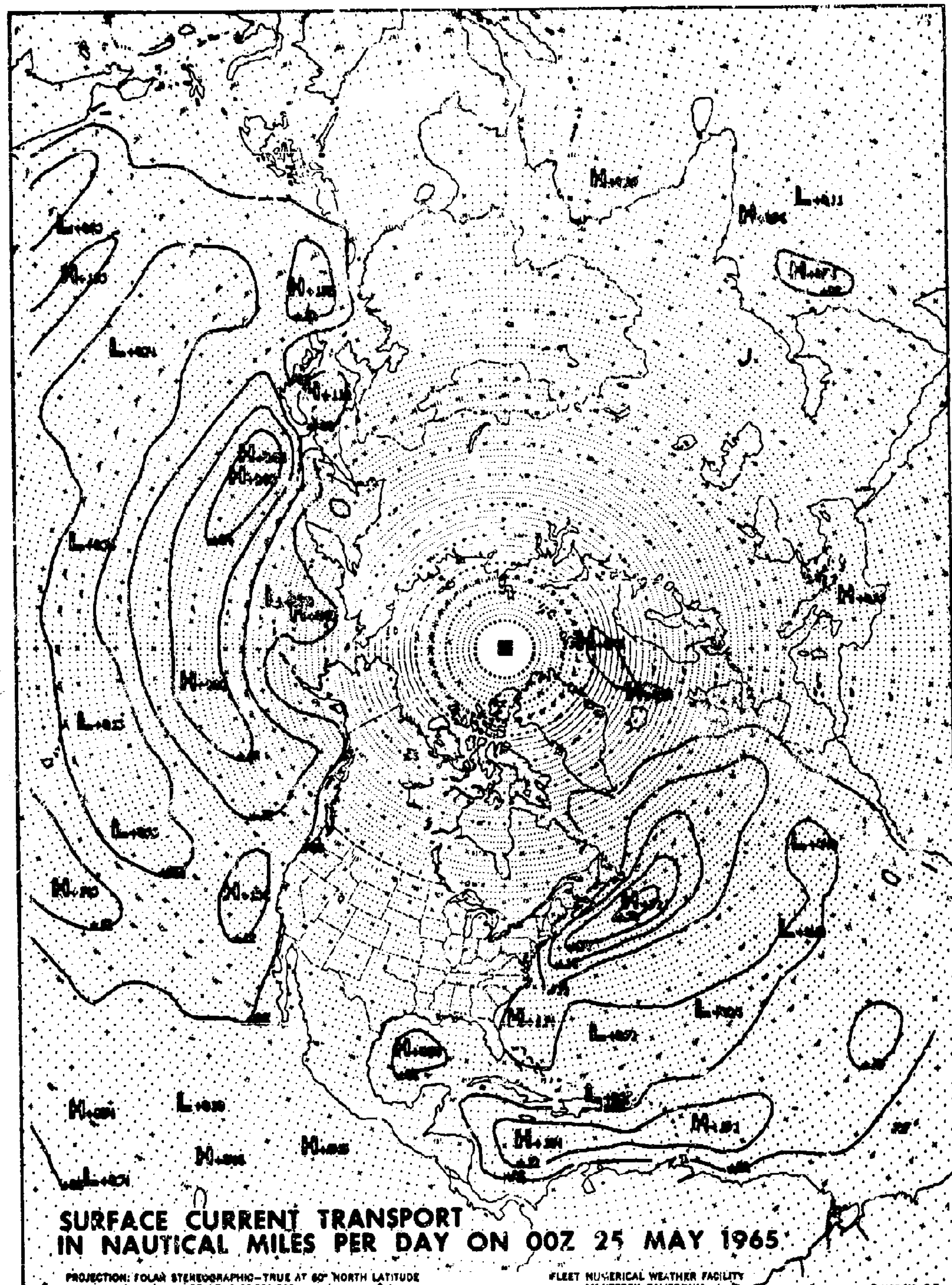

Figure 4.2 Surface Current Transport in Nautical Miles Per Day on 00Z 25 May 1965

Labrador Current, Kuroshio and Cyashio. The low-latitude, westerly return flow (Equatorial current) which results primarily from the "wind component" term is well-defined in both the Atlantic and Pacific. A narrow equatorial countercurrent was obtained as a result of the 200 meter temperature structure used in the characteristic component.

The charts are drawn automatically on an incremental x-y curve plotter. Each chart requires approximately one minute to complete and is of sufficient quality that it can be used immediately for radio-facsimile transmission.

In order to obtain a single continuous field display of both direction and speed of the computed currents, a stream function (ψ) analysis is made, using methods similar to those employed by Bedient and Vederman [7], to present atmospheric flow in the tropics. The vorticity of the current flow is determined from the (u, v) component fields and the Poisson equation.

$$\nabla^2 \psi = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \qquad (26)$$

is solved for ψ using relaxation techniques.

The stream function fields which correspond to the current transport charts on figures 4.1 and 4.2 are shown in figures 4.3 and 4.4. If current vectors are plotted on these charts, there is a very close correspondence, but slight cross contour flow can be observed. This is natural because the derived stream function is nondivergent, while there is divergence in the initial velocity field. In general, the divergence is small in most places, and the stream field provides a good representation of the current pattern.

It can be seen that the hemispheric grid size is too large to show details in areas of strong currents. Detailed models are under development to remedy this shortcoming in areas of relatively dense synoptic weather reports. Some of these models will also include inertia effects and tidal currents. The results of these numerical analyses/forecasts are used as inputs in a number of other oceanographic analyses/forecasts.

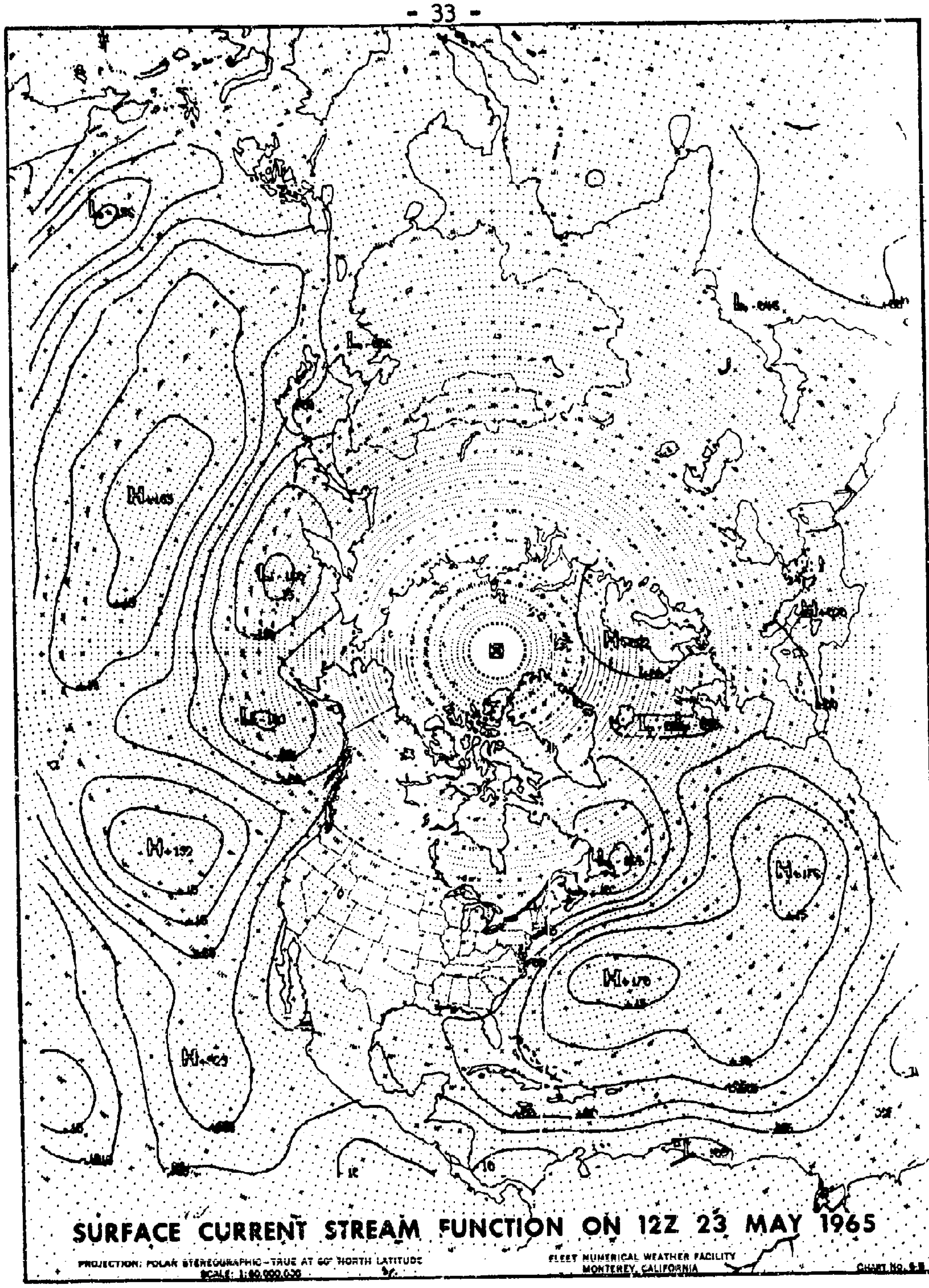

Figure 4.8 Surface Current Stream Function on 12Z 23 May 1965

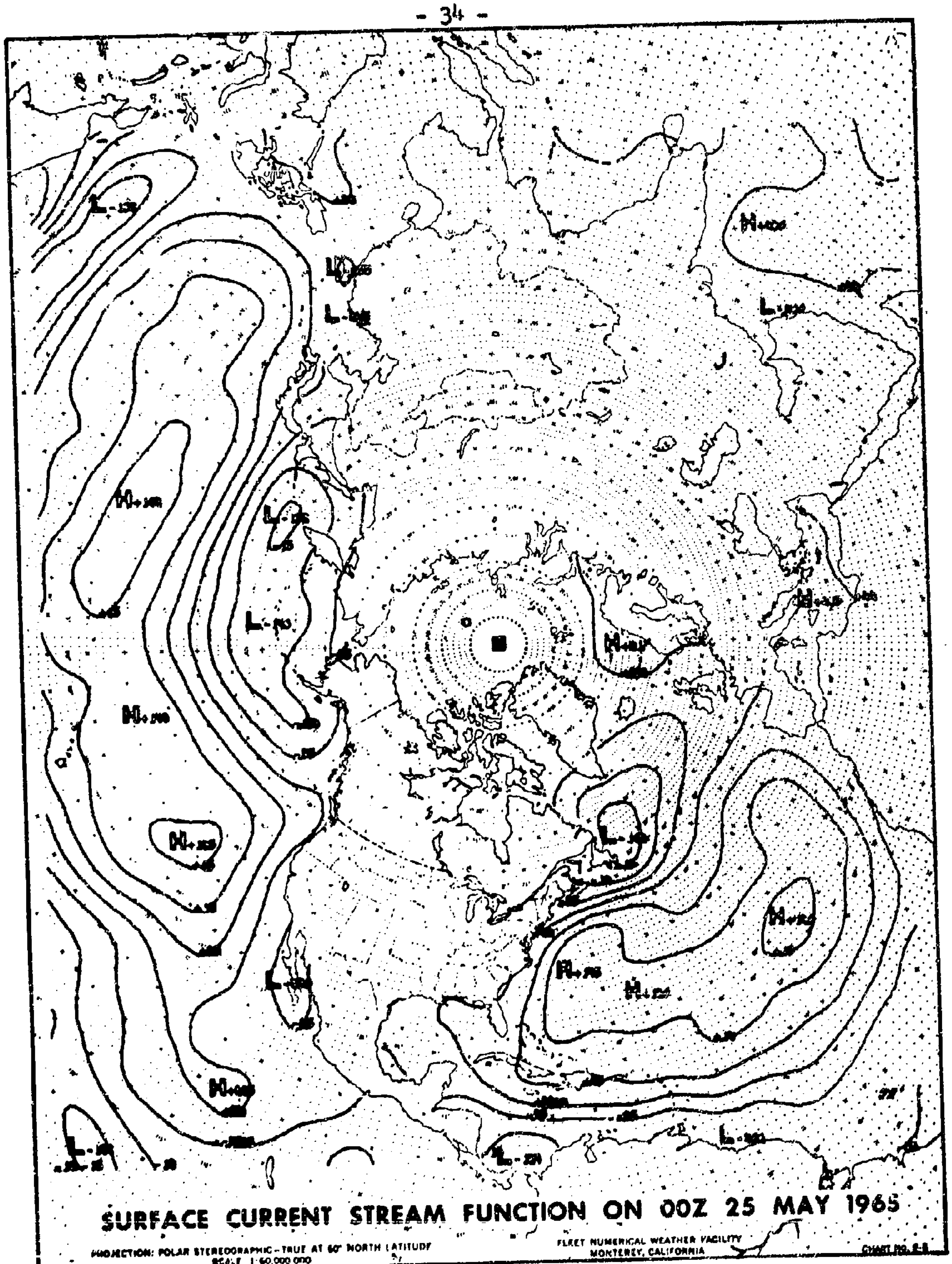

Figure 4.4 Surface Current Stream Function on 00Z 25 May 1965

5. VERIFICATION OF SURFACE CURRENT ANALYSES

A synoptic current chart would be of little value if it could not be verified and the computational scheme tuned as required. Direct current measurements in the open ocean are few, and drift calculations made from navigational fixes are frequently inaccurate in weak current areas, so that it is difficult make a direct evaluation. It has been necessary, therefore, to resort to indirect means which are susceptible to verification on a synoptic basis.

Sea-surface temperature (SST) is the only oceanographic element which permits a reasonably complete synoptic analysis on a hemispheric scale. Such analyses are made twice daily at Fleet Numerical Weather Facility, Monterey (Wolff [128]), and their resolution is such that SST changes can be determined for periods of 24, 48 hours, etc. From these changes are subtracted the local changes computed from air/sea heat exchange equations. If the remainder correlates well with the advective change indicated by $W_{xyzt} \cdot \nabla$ SST the computed currents can be assumed to be reasonably correct.

Some experimental verification has been carried out at Fleet Numerical Weather Facility. The results have been very encouraging. These first computations indicate that the advectional change of SST is, over major parts of the ocean, $< 0.1\,C/(24h)$. In only a few areas does the change rise above $0.3°\,C/(24h)$; higher values are confined to sharp boundaries of major currents. Some discrepancies between analyzed and computed quantities have also been ascertained; these have led to plans of shortening the SST analysis period to decrease the apparent lag. Both SST advection and heat exchange will be used in SST forecasts. Furthermore, 100 nautical miles and smaller mesh lengths will be used in the future for current analyses to make them directly compatible to the SST analyses which are made on a 100 nautical mile grid.

6. THE RESPONSE TIME OF SURFACE LAYERS TO TRANSFER OF MECHANICAL ENERGY

As the surface layer (which is set in motion by wind) has considerable inertia, it can be expected that the response of the sea to the change of driving force is gradual, and that there is a time lag between the change of wind speed and direction and the response of the surface current to this change. A lag also exists in the time for growth of a fully arisen sea in relation to fetch length and duration of the wind. The available data do not allow an exact numerical evaluation of the response time but only a qualitative analysis. These data indicate that the response time for currents varies with the wind speed (especially with the ratio between previous and new [actual] wind speed), the depth of the water, distance from and configuration of the coast. Krause (personal communication) has also found a rapid response in the relatively shallow Kiel Bight.

A few additional observations from areas with slow permanent flow indicate that the complete response time with moderate winds is about three hours. With strong winds and in areas of strong currents, the response is gradual, and the establishment of a steady state takes a longer time (up to 12 hours).

It should be underlined here that the current direction and speed in a given position in the open ocean might not be directly determined by winds prevailing at this position but by stronger, perhaps more persistent and larger wind fields some distance away from the given position.

Hafner [33] offers the following empirical formula for estimation of complete response time in shallow water.

$$\text{Response time (h)} = \frac{\text{Fetch length (km)}}{\sqrt{g \times \text{depth of water (m)}}}$$

This formula would merit testing in deep water, if we assume the depth to be the depth of the surface mixed layer.

It can be expected that verification attempts with numerical analyses/forecasts will shed more quantitative light on the response time.

7. USE OF CURRENT FORECASTS IN NAVAL PROBLEMS; NAVIGATION AND FISHERIES

The analysis and forecasting of surface currents on a synoptic schedule were initiated at Fleet Numerical Weather Facility primarily because of their importance in the following problems:

(a) Quantitative computation of divergence and convergence and the accompanying up and down movement of thermocline depth;
(b) Computation of the decay of transients where one of the prime factors is current;
(c) Forecasting the advective part of sea surface temperature changes;
(d) Construction of detailed dynamic models of thermal structure in areas of strong currents.

Current analyses and forecasts also have other applications. At present, navigators use information on monthly average currents available in atlases and information on tidal currents given on sea charts, in tidal current tables and in pilot books. In many instances, however, it seems to be economically profitable to use prognoses of actual currents, especially in sea and weather routing of ships. This is now being done empirically and indirectly.

At present, allowance is seldom made in the routing practice for drift caused by currents, and, therefore, some inaccuracies might be experienced. In ship routing practice, the resultant of wind currents and permanent flow could be used. The influence of "wave currents" could also be considered as one or two separate factors: (a) involuntary speed reduction and (b) resistance by wave currents. Direct use of information on currents is made in rescue operations, in estimations of drifts of polluted or otherwise contaminated waters and in the estimation of ice flow.

In almost all types of fisheries work there is a need to know the actual currents and to adjust fishing operations accordingly. The need for and the use of currents in fisheries has been summarized by Hela and Laevastu [40]. Current boundaries are extremely important to fisheries. At divergences the deeper, nutrient-rich water is brought into the surface layers where it causes a higher production of organic matter and an aggregation of fish; convergences cause (dynamically) a concentration of zooplankton and are accompanied also by a concentration of fish. Pictorially, one can say that there is an accumulation of everything from plankton to fishermen on a convergence.

Numerical programs for analysis and prediction of current and water type boundaries at the surface are under development and testing at FNWF. These utilize SST gradients as well as surface current gradients.

Past knowledge of actual surface currents has been shaky indeed; this fact has prevented their effective economical use in navigation and fisheries work. The development of good, synoptic analyses and forecasts of ocean currents would benefit many.

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APPENDIX

SYMBOLS AND UNITS USED

A Amplitude of the tide

A_c Current component caused by the change of atmospheric pressure

B_c Breadth of current

C Wave speed

ΔD Dynamic height difference (in dynamic cm.)

d Depth of water

f Coriolis parameter

G_c Permanent flow (gradient current thermohaline and/or characteristic current large-scale, wind-driven)

g Acceleration of gravity

ΔH_g Geostrophic potential ($\beta - \alpha_a \rho_h$)

ΔH_ℓ Longitudinal height difference (of sea level)

ΔH_t Transversal height difference (of sea level)

ΔH_w Slope of sea surface (cm./100 km.)

$H_{1/3}$ Wave height (usually significant height)

h_1 Thickness of top layer

h_2 Thickness of lower (bottom layer)

I Velocity and directional component of surface current caused by influencing factors

i Inclination of the sea surface

K_1 K_2 Constant, factor

$|K$ Unit vector

$L_{1/3}$ Wave length

l Distance between stations;
Distance in nautical miles;
Length of basin

m Beach slope

p Pressure

P_i Periodic portion of inertia current

P_t Periodic portion of tidal current

r Radius of inertia

T Wave period

$\overline{T}$ Mean temperature above zero current

T_p Period of inertia

T_{sfc} Sea-surface temperature

t_i Period of internal waves

V Wind speed (m./sec.)

V_i Velocity of internal waves

W Current speed

W_c Wind current

W_ℓ Longitudinal velocity of current

W_m Mass transport velocity ("wave current")

W_{ptz} Velocity of surface current, with given direction (place, time and depth of the actual current being specified)

W_1; W_2 Relative current speeds in two different layers

x Friction constant (bottom friction; 0.0025 - 0.3 usually taken 0.3)

Δz Depth of zero current

α_σ Specific volume anomaly (constant along σ_t surface)

$\sin \alpha$ Slope of the bottom

β Anomaly of dynamic height

β Deflection of the wind current (in °)

$\tan \gamma$ Downward inclination of the pycnocline

ϕ Average geographical latitude

θ_b Breaker crest angle

ω Angular velocity of the earth (7.29 10 rad./sec.)

$\rho_1 \rho_2$ Densities of two layers (of top and lower layers)

$\Delta\rho$ Density difference at interface

τ Shearing stress of wind

σ_t Sea water density temperature

ψ Stream function

Navy Weather Research Facility (NWRF 36-0667-127)
SYNOPTIC ANALYSIS AND FORECASTING OF SURFACE CURRENTS (A Reprint of Fleet Numerical Weather Facility Technical Note No. 9). June 1967. 55 p., including 15 figures, and 1 appendix.

UNCLASSIFIED

This report describes various techniques for the operational analysis and forecasting of surface currents on a synoptic time scale. It presents a discussion of the general nature and variability of surface currents and evaluates separately the various components composing the surface currents, as well as their modifying factors. Suggestions are given for the use of these current forecasts in naval problems, navigation and fisheries.

Results of large scale, synoptic current calculations are shown.

1. Oceanography
2. Currents
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